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Neotectonics of the upper Mississippi embayment

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Abstract

Although the upper Mississippi embayment is an area of low relief, the region has been subjected to tectonic influence throughout its history and continues to be so today. Tectonic activity can be recognized through seismicity patterns and geological indicators of activity, either those as a direct result of earthquakes, or longer term geomorphic, structural, and sedimentological signatures. The rate of seismic activity in the upper Mississippi embayment is generally lower than at the margins of tectonic plates; the embayment, however, is the most seismically active region east of the Rocky Mountains, with activity concentrated in the New Madrid seismic zone. This zone produced the very large New Madrid earthquakes of 1811 and 1812.

Geological and geophysical evidence of neotectonic activity in the upper Mississippi embayment includes faulting in the Benton Hills and Thebes Gap in Missouri, paleoliquefaction in the Western Lowlands of Missouri, subsurface faulting beneath and tilting of Crowley's Ridge in northeastern Arkansas and southeastern Missouri, subsurface faulting along the Crittenden County fault zone near Memphis, Tennessee, faulting along the east flank of the Tiptonville dome, and numerous indicators of historic and prehistoric large earthquakes in the New Madrid seismic zone.

Paleoearthquake studies in the New Madrid seismic zone have used trenching, seismic reflection, shallow coring, pedology, geomorphology, archaeology, and dendrochronology to identify and date faulting, deposits of liquefied sand, and areas of uplift and subsidence. The cause of today's relatively high rate of tectonic activity in the Mississippi embayment remains elusive. It is also not clear whether this activity rate is a short term phenomenon or has been constant over millions of years. Ongoing geodetic and geological studies should provide more insight as to the precise manner in which crustal strain is accumulating, and perhaps allow improved regional neotectonic models.

1. Introduction

It is easy to think of the Mississippi River and it's surrounding expanse of nearly level flood plains as evidence of a region that is tectonically and seismically dead. However, throughout its history, the Mississippi River has been profoundly influenced by tectonics and, indeed, owes its very existence to tectonics. The Mississippi River flows in an ancient trough of tectonic origin, as is the case with most of the world's major rivers (Potter, 1978). This trough has existed in one form or another since at least early Cambrian time, when the Iapetus (proto-Atlantic) Ocean opened, and a

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rift in the Earth's crust, the Reelfoot rift, formed nearly coincident with the current Lower Mississippi Valley (LMV) (Braile et al., 1982). Clearly, the neotectonic activity that we describe in this paper is not great, or has only recently begun. If this were not true then more significant topography *would* be evident (Schweig and Ellis, 1994).

A discussion of neotectonic activity in the LMV can be conveniently framed in terms of genesis and geography into three major categories: growth faults of the Gulf Coastal Plain, regional warping in Louisiana, southeastern Arkansas, and Mississippi, and seismicity and deformation in the upper Mississippi embayment, particularly the New Madrid seismic zone (NMSZ). Because the first two categories are discussed in more detail elsewhere in this volume (Roberts and Coleman, this volume), we will concentrate on the evidence for neotectonics in the upper Mississippi embayment.

H.N. Fisk, to whom this special volume is dedicated, recognized regional faulting and effects of the New Madrid earthquakes of 1811–1812; however he did not seriously investigate the effects of tectonics on the Mississippi River or prehistoric earthquakes. His work, of course, predates most of the modern ideas and methods of neotectonic investigation. We therefore will not discuss his contributions in this paper, but the reader is encouraged to read the Appendix of Fisk (1944), which does discuss the major structural features of the LMV as well as the New Madrid earthquakes.

2. Neotectonics of the Upper Mississippi embayment

A reasonable first question that one might ask is how do we even recognize that active tectonics is occurring in an area such as the LMV, and in particular the Mississippi embayment, where we have no large fault scarps or mountains? Tools available to us include seismicity patterns and geological indicators of activity, either those as a direct result of earthquakes, or a longer term geomorphic and sedimentological signature. Geophysical techniques such as gravity, magnetics, magnetotellurics, seismic refraction, and seismic reflection may also allow association of tectonic activity with particular geological structures and give clues as to the causes of that activity (Hildenbrand et al., 1995).

2.1. Seismological evidence for neotectonic activity

It would generally be agreed that a first order indicator of tectonic activity in any area would be seismicity. In terms of seismic hazard analysis, the entire LMV is in what has been referred to as the North American stable continental region. This is the part of the interior of the North American plate that has not undergone any major orogenic activity since the Early Cretaceous (Kanter, 1994). As one might expect, seismicity in such regions is generally lower than at the margins of tectonic plates. Fig. 1 shows the seismicity of the central United States from 1811 to 1987. Clearly the most dramatic feature in the region is the NMSZ, stretching about 250 km from west of Memphis into southern Illinois. This zone produced the very large New Madrid earthquakes of 1811 and 1812 and remains the most active seismic zone east of the Rocky Mountains.

2.1.1. The New Madrid earthquakes of 1811–1812

The New Madrid earthquake sequence began on December 16, 1811 and continued for at least a year. The sequence is notable in apparently containing three large main shocks, each with a moment magnitude, **M**, of about 8 (Johnston, 1996). This would make these earthquakes the largest historically documented earthquakes to have occurred in a stable continental interior anywhere on Earth and the largest earthquakes to have struck the conterminous United States in recorded history. Because seismic energy attenuates slowly in the central US, the New Madrid earthquakes may be the strongest historical earthquakes in recorded history in terms of area affected (Johnston and Kanter, 1990). For example, these three earthquakes had felt areas of at least 5 000 000 km^2, and possibly significantly more (Nuttli, 1982; Stover and Coffman, 1993). They were felt as far away as Boston, Massachusetts

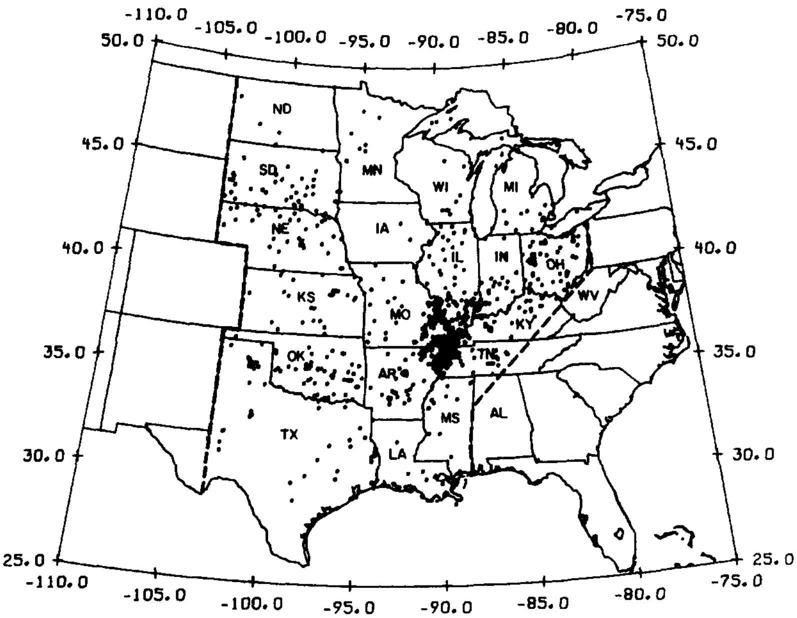

Fig. 1. Map of earthquakes with body-wave magnitude, m_b, 3.0 in the central US (area within heavy dashed lines) between 1811 and 1987 (after Mitchell et al., 1991).

(1690 km). In addition to the main shocks, thousands of aftershocks were associated with the New Madrid earthquake sequence, many of which caused damage and were felt along the eastern seaboard (Street and Nuttli, 1984).

Although the exact epicenters of the 1811–1812 New Madrid earthquakes are unknown, their relationship to the NMSZ is strongly suggested by the isoseismal maps compiled from historical accounts (Nuttli, 1973) and the distribution of ground failure, particularly liquefaction (Fuller, 1912; Obermeier, 1984) and earthquake-induced landslides (Jibson and Keefer, 1988,1989). The first large New Madrid earthquake likely occurred on the southernmost trend of seismicity shown on Fig. 2, the second shock on the northernmost trend, and the third on the central northwest-trending segment (Johnston and Schweig, 1996).

The total area affected by ground failure in 1811–1812, including fissures, sandblows, landslides, and subsidence was about 48 000 km^2 (Street and Nuttli, 1984). In the epicentral region, uplift and subsidence on the order of a couple of meters occurred over hundreds of square kilometers (Fuller, 1912; Russ, 1982). Liquefaction of subsurface sand layers ejected sand, water, and other materials through fissures, some several kilometers long and many tens of meters wide (Fuller, 1912). Saucier (1977) has estimated that 10 500 km^2 were inundated by as much as one meter of ejected sand and water. Massive bank failures along the Mississippi River transported large tracts of land into the river channel, which was reported to have been choked with trees and the wreckage of boats (Penick, 1981). At two locations, one upstream and one downstream from New Madrid (Penick, 1981; Johnston, 1982; Van Arsdale et al., 1995a), waterfalls or rapids formed in the Mississippi River channel; these disturbances in the soft sediments of the riverbed were eroded and obliterated rapidly. Landslides along the bluffs bordering the Mississippi River alluvial valley occurred from near Cairo, Illinois, at least to near Memphis (Fuller, 1912; Jibson and Keefer, 1988). Eyewitnesses describe the epicentral region land surface as being so disrupted that in many places it was uninhabitable (Penick, 1981).

2.1.2. Other notable historical earthquakes

Since 1812 at least 28 damaging earthquakes having moment magnitudes between 4.1 and 6.8

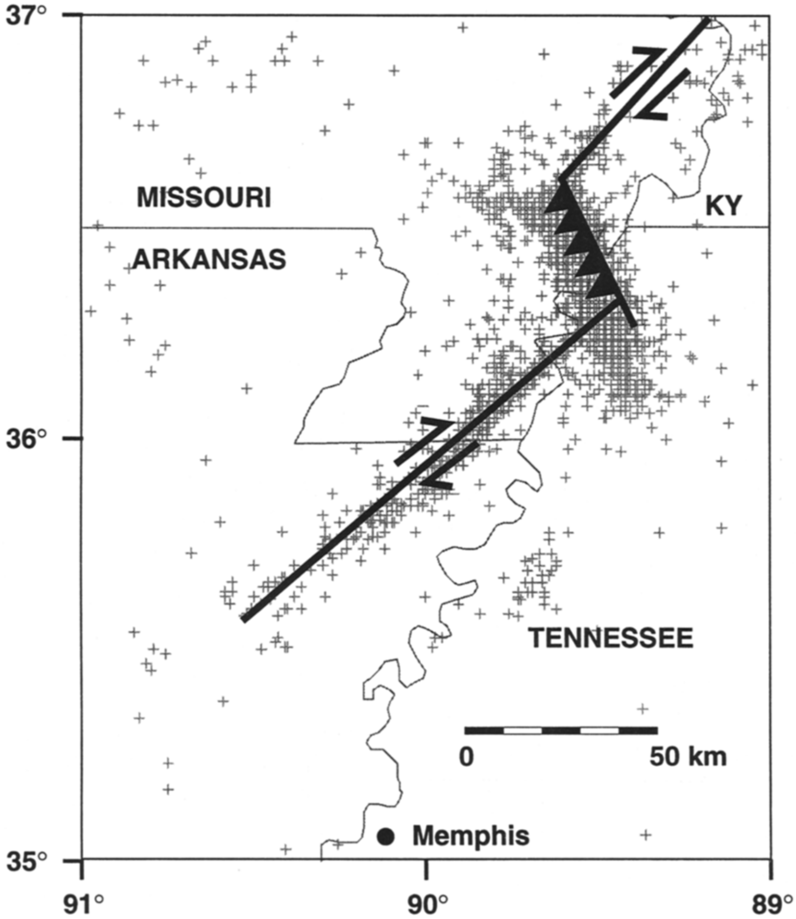

Fig. 2. Earthquakes of the NMSZ. Gray crosses indicate locations of recent (1974–1991) seismicity. Bold lines show simple tectonic model of the seismic zone as a left-stepping, right-lateral strike-slip fault zone (Russ, 1982).

have struck the New Madrid region (Nuttli, 1983; Hamilton and Johnston, 1990). The two largest of these occurred near the southern and northern ends of the NMSZ. A $\mathbf{M}=6.4$ earthquake occurred near Marked Tree, Arkansas, in 1843. It was felt over an area of about 1 500 000 km^2 (Nuttli, 1974) and caused severe damage at what is now Memphis (Stover and Coffman, 1993).

At the northern end of the seismic zone, a $\mathbf{M}=6.8$ earthquake occurred near Charleston, Missouri, in 1895, the largest to have struck the region since the 1811–1812 earthquake sequence. The earthquake was felt over 2 500 000 km^2 (Nuttli, 1974), from Ontario, Canada, to New Orleans, Louisiana. Damage was severe in Charleston and significant in Cairo. Liquefaction of sand and eruption of liquefied sand onto the ground surface was extensive and occurred over a region about 16 km in diameter (Obermeier, 1988).

2.1.3. *Microseismicity*

The instrumental seismic record of the New Madrid region is shown in detail in Fig. 2. It shows that modern microseismicity is concentrated in several linear bands collectively called the NMSZ. Two of these bands trend north-northeast and are connected by a northwest-trending band of seismicity. The two north–northeast trends of seismicity do not correlate with any known faults that intersect the surface. Seismic-zone focal mechanisms indicate that right-lateral strike-slip is occurring along these trends (Herrmann and Canas, 1978; O'Connell et al., 1982) and a study of well located epicenters shows that they occur along nearly vertical fault planes (Chiu et al., 1992). The northwest-trending seismicity zone appears to be occurring along a thrust fault that daylights along the Reelfoot scarp discussed below. Thus, the seismicity can be thought of as a right-lateral strike-slip fault system connected by a left-stepping thrust fault, as proposed by Russ (1982).

2.1.4. Rates of recurrence of earthquakes

How often such large earthquakes strike the LMV is not only critical to seismic hazard analysis, but to the interpretation of the neotectonics of the region. A $\mathbf{M}=8$ earthquake may produce 8–10 m of slip. Repeated earthquakes of such a magnitude would clearly result in disruption of fluvial systems, as well as the landscape in general. Seismological, geodetic, and most paleoseismological data suggest a surprisingly short recurrence interval in the NMSZ, on the order of a thousand years or less, with deformation rates comparable to those at plate margins. Yet other data, particularly the lack of topography, indicate that the rapid rates of crustal strain implied by such intervals cannot have been occurring for geologically long periods of time.

One line of evidence for a short recurrence interval for large earthquakes is in the form of earthquake frequency–magnitude relationships. Johnston and Nava (1985) extrapolated the historical and instrumental record (mostly small earthquakes) and determined that earthquakes similar in magnitude to the 1811–1812 New Madrid earthquakes should recur every 550–1100 years on average in the NMSZ.

Liu et al. (1992) reoccupied a 1950s triangulation network in the southern NMSZ using the Global Positioning System (GPS). Their data indicate unexpectedly rapid crustal shear strain accumulation of the order of 10^{-7}/year, which results in 5–7 mm/year of right-lateral slip over the width of the network. At this rate of deformation, enough strain energy to produce an 1811–1812-type event could accumulate in 400–1100 years (Schweig and Ellis, 1994).

Paleoseismological studies, which will be discussed below, indicate similarly short or shorter recurrence intervals for earthquakes large enough to cause liquefaction over a significant portion of the NMSZ.

Other types of data argue against such a short interval or, alternatively, that such high rates must be a short-term phenomenon (Schweig and Ellis, 1994). A $\mathbf{M}=8.0$ New Madrid earthquake can be expected to produce about 8 m of slip. Evidence of such high rates of slip (9 km m^{-1} year^{-1}) are lacking. Schweig and Ellis (1994) argue that, even if the slip is distributed across the late Precambrian to early Paleozoic Reelfoot rift (Fig. 3), which bounds much of the New Madrid seismicity, such large deformation should have a significant topographic signature and should have left a clear record in the near-surface stratigraphic section. Such a record of extensive deformation is not observed in the topography or at the top of the Cretaceous section that is clearly imaged in high-resolution seismic reflection profiles of the subsurface of the NMSZ (e.g., Luzietti et al., 1995; Schweig et al., 1992b; Sexton et al., 1992; Van Arsdale et al., 1995b). In addition, aeromagnetic and gravity surveys suggest that several Paleozoic or older upper crustal features that cross the northern trend of New Madrid seismicity have cumulative offsets of less than 10 km (Hildenbrand and Hendricks, 1995).

Within the LMV, but outside of the NMSZ, microseismicity levels are much lower and more diffuse. There are subtle linear trends of earthquake epicenters along the margins of Reelfoot rift and, perhaps, the Bootheel lineament (Fig. 3), although the significance of this is unknown.

2.2. Geological and geophysical evidence for neotectonic activity

Geological and geophysical (gravity, magnetics, seismic reflection) evidence for neotectonic activity has been accumulating rapidly over the past two decades, and particularly in the past 6 years. Paleoseismology, that is the study of prehistoric earthquakes through their geological effects, has been of particular benefit. The sites examined have been either in backhoe trenches, drainage ditches, or shallow cores. The geological features indicative

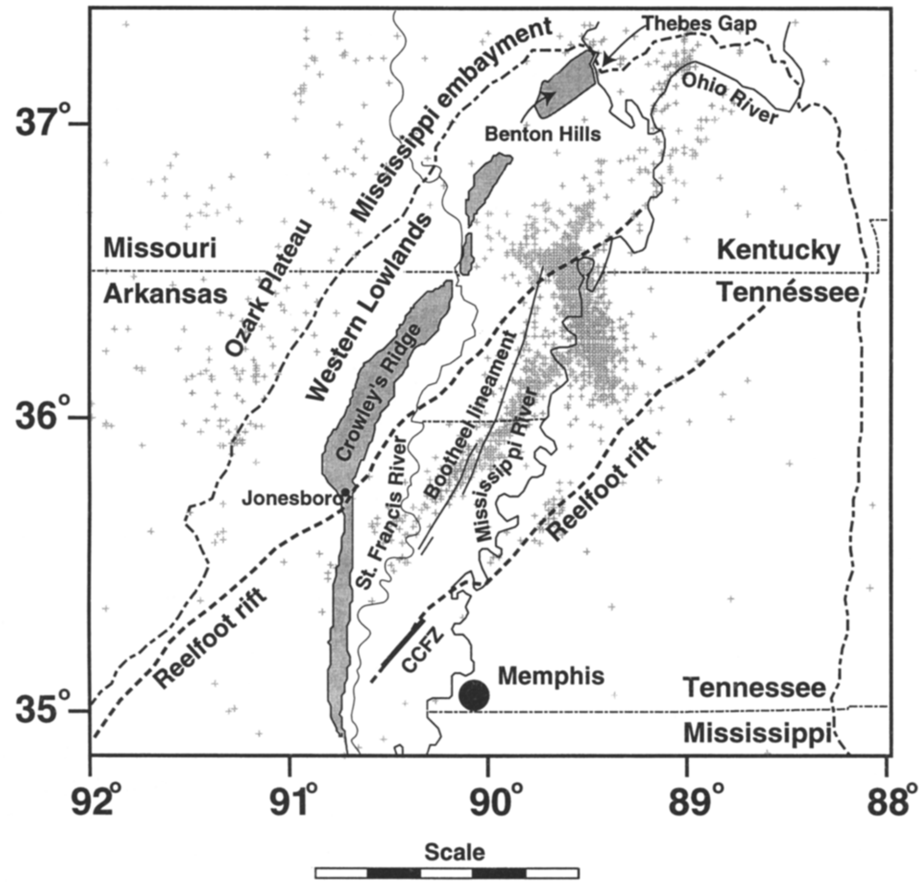

Fig. 3. Tectonic elements of upper Mississippi embayment discussed in text. Sesimicity as in Fig. 2. CCFZ, Crittenden County fault zone.

of earthquakes include liquefaction of subsurface sands, faulting, and warping of the ground surface. Herein, we present evidence of tectonic activity by geographic area, starting in the Thebes Gap area, working southward along the west side of the embayment, and back up through the NMSZ. The areas discussed are indicated in Fig. 3.

2.2.1. Northern Crowley's Ridge: Benton Hills-Thebes Gap area, Missouri and Illinois

Thebes Gap of Missouri and Illinois, lies along a narrow reach of the Mississippi River that is bounded by the Santa Fe Hills on the north and the Benton Hills on the south. These hills are portions of the 320-km-long Crowley's Ridge that extends from Thebes, Illinois, to Helena, Arkansas (Fig. 3). Dissection in the gap has exposed structurally deformed sedimentary strata that range in age from Ordovician through the Miocene to Pliocene Mounds gravel (McQueen, 1939; Stewart, 1942; Stewart and McManamy, 1944; Pryor and Ross, 1962; Johnson, 1985; Harrison and Schultz, 1994). Harrison and Schultz (1994) discuss the structural history of Thebes Gap and note the extensive post-Mounds Gravel deformation occurring along reactivated Paleozoic faults. The faults predominantly strike northeast, are vertical, and post-Mounds movement consists of minor reverse displacement and major right-lateral strike-slip offset. Most of the faulting is believed to have occurred over 200 000 years ago (Harrison and Schultz, 1994).

A major fault in the Thebes Gap area is the Commerce fault. This fault strikes northeast and extends along the eastern margin of the Santa Fe Hills in Illinois, and the Benton Hills in Missouri, and parallels the northwestern margin of the Reelfoot rift. The fault shows up dramatically in regional aeromagnetic studies, but appears to displace other magnetic anomalies no more than 5 km laterally (Hildenbrand and Hendricks, 1995). The southwestern extension of the Commerce fault coincides with the Idalia Hill fault in Missouri (Stewart, 1942; Grohskopf, 1955). Although when their paper was written there was no definitive evidence for Holocene faulting in the Thebes Gap area, Harrison and Schultz (1994) note that traces of the English Hills Albrecht Creek, Rock Springs, and Simpson's Rocks faults, which splay off of the Commerce fault, are visible across areas covered by loess. They further note that there is anomalous thickening of Mississippi River alluvium between known faults. Recent excavations across the English Hill fault in the Benton Hills have uncovered Peoria loess faulted against Plio-Pleistocene Mounds gravel (D. Hoffman and R. Harrison, personal communication, 1995). Moderate seismicity in the Thebes Gap vicinity has produced historic earthquakes of up to $\mathbf{M}=4.4$.

2.2.2. Western Lowlands in Missouri

Paleoliquefaction has been reported in the Western Lowlands of Missouri, particularly along the St. Francis River (Fig. 3) (Vaughn, 1991, 1992; Vaughn et al., 1993). The Western Lowlands lies between the Ozark Plateau and Crowley's Ridge and is a former valley of the ancestral Mississippi River (Fisk, 1944). Vaughn cites many tens of clastic dikes, two buried sandblow events, and three surficial sandblow events in the Western Lowlands. His estimates of the timing of the prehistoric events are: 23 000–17 000 years B.P., 13 430–9000 years B.P., A.D. 240–1020, and A.D. 1440–1540 (sites marked V on Fig. 4). Vaughn further speculates that a northwest-trending fault may underlie the St. Francis River or that a fault may lie beneath the margin of Crowley's Ridge thus providing a local seismic source for the paleoearthquakes. The proposed fault beneath the margin of Crowley's Ridge may be the southwestern extension of the Commerce fault (Vaughn, 1991; Harrison and Schultz, 1994). Alternatively, this area of liquefaction could represent distal liquefaction from NMSZ paleoearthquakes.

2.2.3. Central Crowley's Ridge near Jonesboro, Arkansas

Crowley's Ridge, of northeastern Arkansas and southeastern Missouri, has historically been explained as being formed by incision of the ancestral Mississippi River flowing along its western side and concurrent incision by the ancestral Ohio River flowing along its eastern side. However, Fisk (1944) and O'Leary and Hildenbrand (1978) proposed local ridge-bounding faults, and Cox's (Cox, 1988) geomorphic analysis of drainage basins on Crowley's Ridge suggests Quaternary southeasterly tilting of the ridge segment north of Jonesboro. Subsequent seismic reflection surveys (Nelson and Zhang, 1991; Van Arsdale et al., 1994b, Van Arsdale et al., 1995b) have revealed that the margins of Crowley's Ridge near Jonesboro are underlain by faults (Fig. 5). The faults are steep, display normal and reverse displacement, and have been interpreted as upper portions of flower structures (Van Arsdale et al., 1995b). These faults locally uplift Paleozoic through Eocene sediments from 30 to 60 m beneath Crowley's Ridge, but near-surface displacement appears to be less than 8 m. Near Jonesboro, Crowley's Ridge has 60 m of topographic relief and thus most of the ridge's relief must be erosional. Based on the distribution of late Pleistocene terraces on either side of the ridge, Van Arsdale et al. (1995b) speculate that Wisconsin uplift of Crowley's Ridge (or arching centered on Crowley's Ridge) shifted the ancestral Mississippi River westward and the ancestral Ohio River eastward thus strongly influencing the denudational history of these two rivers. Trenches have been recently excavated across two margins of Crowley's Ridge (Drouin, 1995). One trench reveals flexed Holocene sediments above a margin fault but it is not clear if the flexing is geomorphic or tectonic.

Seismic reflection data obtained across west-bounding faults of the Reelfoot rift east of Crowley's Ridge also demonstrate recurrent fault displacement (Van Arsdale et al., 1994b, 1995b).

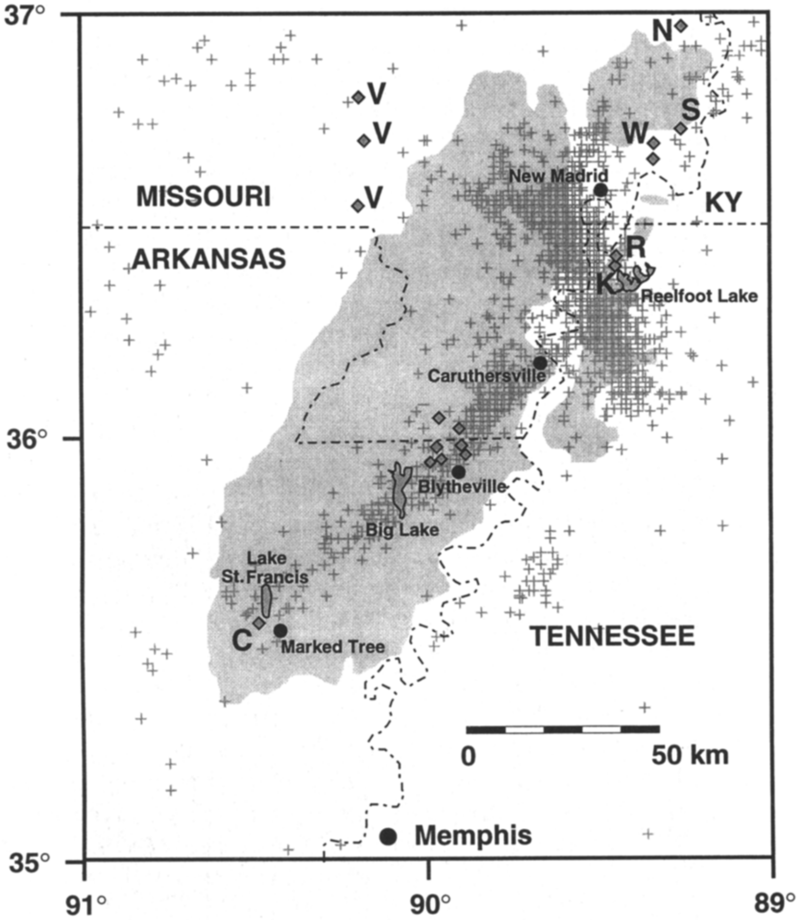

Fig. 4. Sites of NMSZ paleoearthquake sudies. Seismicity as in Fig. 2. Shading represents >1% of area covered by sand-blow deposits (Obermeier, 1989). Gray diamonds indicate sites at drainage ditches or trenches displaying prehistoric earthquakes features; R, Russ (1982); K, Kelson et al. (1992,1996); S, Saucier (1991); and V, Vaughn (1991). N and W described in Li et al. (1994). C and sites near Blytheville decribed in Tuttle and Schweig (1995) and Lafferty et al. (1996).

Ninety meters of post-Cretaceous displacement has occurred across a major down-to-the-east normal fault (Fig. 6). This fault zone has also displaced Quaternary strata approximately 3.5 m at a depth of 35 m. However, the Quaternary displacement appears to be strike-slip and reverse movement. Seismicity beneath Crowley's Ridge and along this portion of the western margin of the Reelfoot rift is very low, but the current stress field is favorably oriented for reactivation of these faults and the very high strain rates measured across the western margin of the Reelfoot rift north of Jonesboro (Liu et al., 1992) suggest that strain may be accumulating along the west-bounding faults of the Reelfoot rift.

2.2.4. The Crittenden County fault zone

The Crittenden County fault zone forms part of the southeastern boundary zone of the Reelfoot rift (Crone, 1992; Luzietti et al., 1995) (Fig. 3). Although currently seismically quiet, the Crittenden County fault zone is just 30 km from Memphis and thus has been of considerable concern as a possible source of future earthquakes. The fault zone is a steeply dipping zone of disruption, with minor reverse faulting and a down-to-

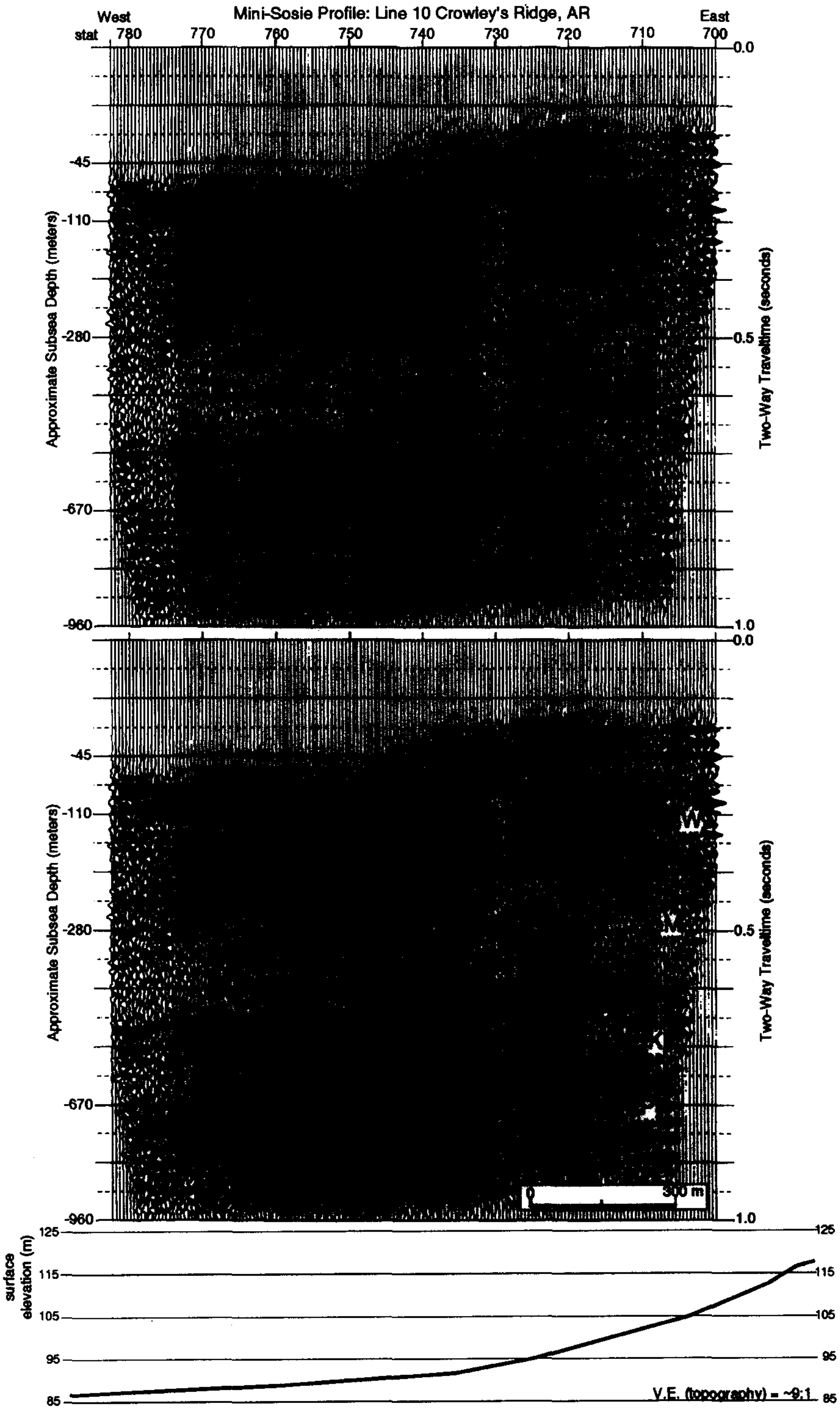

Fig. 5. A 1.2-km-long, east–west seismic reflection line across western margin of Crowley's Ridge, approximately 20 km south of Jonesboro, AR (Fig. 3) (line 10 of Van Arsdale et al., 1995b). Paleozoic through Eocene Wilcox strata have been uplifted beneath ridge. Top, uninterpreted; bottom, interpreted.

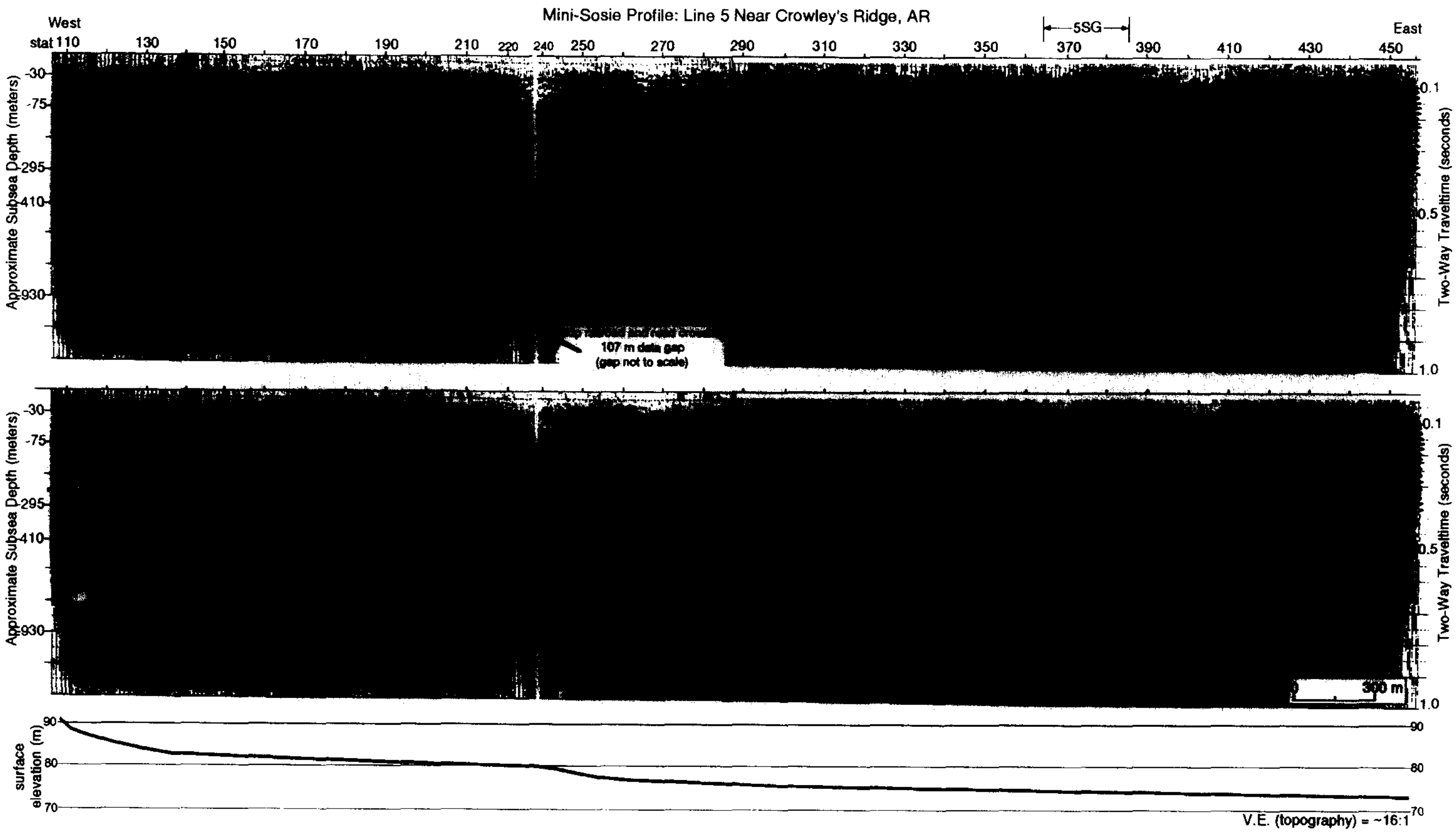

Fig. 6. A 5.4-km-long, east–west seismic reflection line across the eastern margin of Crowley's Ridge (at Sta. 240) and across a west-bounding fault of the Reelfoot rift (Sta. 390). The line is located approximately 7 km northeast of Jonesboro, AR (Fig. 3) (line 5 of Van Arsdale et al., 1995b). Paleozoic though Eocene Wilcox strata are displaced across the Reelfoot rift fault. Top, uninterpreted; bottom, interpreted.

the-southeast monoclinal flexure. The existence of the Crittenden County fault zone was first inferred on the basis of well log data interpreted by Caplan (1954) and further supported by magnetic data (Hildenbrand, 1982). Geophysical well log data collected in the vicinity of the Crittenden County fault zone indicate a maximum of roughly 82 and 63 m of vertical displacement of the Paleozoic and Cretaceous rocks, respectively (Luzietti et al., 1995).

Deep seismic reflection (Vibroseis) lines crossing the Crittenden County fault zone show that the zone has a reverse, down-to-the-southeast sense of displacement at the top of the Cretaceous and Paleozoic strata, and that it has experienced reverse faulting as recently as the late Tertiary (Crone, 1992). Shallow high resolution seismic reflection lines (using an earth tamper energy source) along the Crittenden County fault zone clearly show a maximum of 70 m of structural relief near the Mississippi River, decreasing to the southwest, with a termination of discernible faulting northwest of Memphis (Luzietti et al., 1995; Nicholas et al., 1992).

The age of the youngest deformation along the Crittenden County fault zone is not certain, but some deformation of Quaternary units has been interpreted on the high resolution lines (Nicholas et al., 1992; Luzietti et al., 1995). Additionally several very high resolution seismic reflection surveys (shotgun and jack hammer energy sources, better suited for resolving strata between 5 and 100 m below the surface) have been collected and one of them shows evidence of Holocene faulting at about 6 m depth (Williams et al., 1995). A 2-m-deep trench was dug across this same location revealing flexure is present in strata less than 1 m below the surface, but no obvious faulting was observed (Crone et al., 1995).

2.2.5. *The New Madrid seismic zone*

As discussed above, the NMSZ is clearly the most seismically active, and therefore probably the most tectonically active, area in the LMV region. As such it has been of great concern in terms of regional earthquake hazards and has been a focus area of study for the US. Geological Survey under the National Earthquake Hazards Reduction Program (e.g., Hamilton and Johnston, 1990; Johnston and Shedlock, 1992; Shedlock and Johnston, 1994). In particular, paleoseismological studies are being used to determine how often earthquakes have occurred in the past in the seismic zone. A variety of paleoseismological indicators are being used including faulting, regional deformation, liquefaction deposits, and dendrochronology. These studies are discussed below for each of the three seismicity trends that make up the NMSZ. In addition, elsewhere in this volume, Schumm and Spitz (1996) present a geomorphic evaluation of the northern portion of the NMSZ that may be compared to the studies discussed below.

2.2.5.1. Southern New Madrid seismic zone. The southern arm of the NMSZ stretches from about Marked Tree to Caruthersville, Missouri (Figs. 2–4). It has been noted that this is along the axis of the Reelfoot rift and the epicenters lie above the Blytheville arch, a mostly pre-Late Cretaceous anticlinal feature defined by seismic reflection (Hamilton and McKeown, 1988; McKeown et al., 1990). Extensive surface liquefaction is present along this seismicity trend, which is assumed to reflect the fault that ruptured during the first New Madrid earthquake in 1811 (Johnston, 1996).

Although there is no clear evidence of surface faulting from the 1811–1812 or prehistoric earthquakes along the southern seismicity trend, it has been suggested that an enigmatic feature named the Bootheel lineament may be a coseismic break of the 1811 event (Schweig and Marple, 1991). The lineament, first noted on satellite imagery (Marple and Schweig, 1992; Schweig and Marple, 1991) extends about 135 km, from just east of Marked Tree north-northeastward to west of New Madrid, Missouri (Fig. 3). Although it does not lie directly on the southern seismicity trend, it does cut the seismicity trend at a low angle near Blytheville, Arkansas. The nature of the trace of the Bootheel lineament varies along strike but is generally represented by shallow linear depressions, commonly containing standing water and continuous or discontinuous linear bodies of sand. Trenches excavated across the lineament indicate that the sand bodies are the surface expression of dikes of liquefied sand. Dating suggests that this sand was likely ejected during the 1811–1812 earthquakes (Schweig and Marple, 1991; Schweig et al., 1992a; Crone et al., 1995).

The detailed pattern of the Bootheel lineament traces are very similar in geometry to a right-lateral strike-slip fault and seismic reflection profiles have been interpreted to show a complex zone of deformation consisting of multiple flower structures and fractured rock, with deformation at least as young as the base of the Quaternary (Schweig et al., 1992b).

Although the origin of the Bootheel lineament remains enigmatic, a number of "sunk lands" (Fuller, 1912) lie above the northwestern flank of the Blytheville arch and appear to be tectonic in origin. Two large sunklands in northeastern Arkansas are Big Lake and Lake St. Francis (Fig. 4). Saucier (1970) believed that these lakes were formed due to ponding from the growth of natural levees; however, King (1978) and Guccione and co-workers (Guccione et al., 1988,1993,1994; Guccione and Van Arsdale, 1995) believe that these lakes and associated sunklands owe their present form to 1811–1812 deformation. Big Lake appears to have formed by subsidence and downstream uplift along the south-flowing Little River. Coring within Big Lake reveals two organic mats that apparently reflect 1811–1812 subsidence and a prehistoric subsidence (possibly co-seismic) event (Guccione et al., 1993; Guccione and Van Arsdale, 1995). Similarly, at Lake St. Francis, the St. Francis River has been locally ponded by subsidence and downstream uplift. At Lake St. Francis, however, there is core data to support four ponding events in the last 8000 years (Guccione and Van Arsdale, 1995).

Dendrochronological studies of baldcypress trees at Big Lake and Lake St. Francis have been undertaken to assess the seismic histories of these areas (Van Arsdale et al., 1994a). Apparently due to timbering, no baldcypress trees that predate 1811 were found at Big Lake; however, numerous baldcypress at Lake St. Francis are older than 1811 and they experienced a major growth suppression between 1813 and 1840. Of particular interest is that the dendrochronological record at Lake St. Francis extends to A.D. 1321 and the only major tree ring growth anomaly is the 1813–1840 growth reduction (Cleveland and Stahle, 1994). Thus, the dendrochronological record at Lake St. Francis suggests that no great earthquakes affected the lake in the 490 years prior to 1811.

Most of the liquefaction evidence examined to date for recurrent earthquakes in the southern NMSZ comes from the area near Blytheville with additional evidence at the southern end of the seismic zone near Marked Tree (Fig. 4). In the NMSZ, surficial sand blows are so large (commonly 1.0–1.5 m in thickness and 10–30 m in diameter) that they have been only slightly modified by plowing and are still easy to identify on aerial photographs and on the ground. Dating of these liquefaction features requires a multidisciplinary approach involving soil profile development, radiometric dating, and archaeology (Tuttle and Schweig, 1995; Lafferty et al., 1996).

At Marked Tree (site C on Fig. 4), three liquefaction episodes are recorded, with the youngest assumed to be from the 1811–1812 earthquake sequence. The oldest is archaeologically constrained to date from A.D. 0 to A.D. 500 and a third deposit is constrained to date from between A.D. 1600 and A.D. 1811 (Lafferty et al., 1996).

At least seven sites in the region between Blytheville and Caruthersville contain prehistoric liquefaction (sites near Blytheville on Fig. 4). This region not only experienced intense liquefaction in 1811–1812 and earlier earthquakes, but has a rich and well-preserved archeological history (e.g., Morse and Morse, 1983). For the sites dated thus far, there appear to be three liquefaction events that predate the 1811–1812 earthquakes. Estimates of the ages are A.D. 0 to A.D. 500, A.D. 800 to A.D. 1000, and A.D. 1200 to A.D. 1400 (Lafferty et al., 1996).

2.2.5.2. Central New Madrid seismic zone. Numerous studies have documented the tectonic origin of landforms in the central portion of the NMSZ (Fuller, 1912). The most obvious tectonic landforms in this region are the Reelfoot Lake basin, Lake County uplift, and the Reelfoot scarp (Fuller, 1912; Krinitzski, 1950; Stearns, 1979, Crone and Brockman, 1982; Hamilton and

Zoback, 1982; Russ, 1982; Kelson et al., 1992, 1996; Stahle et al., 1992; Merritts and Hesterberg, 1994) (Fig. 3). A tremendous post-1812 growth surge revealed in baldcypress tree rings in Reelfoot Lake supports the interpretation that Reelfoot Lake formed during 1812 co-seismic uplift of the Lake County uplift and coincident ponding of the west-flowing Reel Foot River (Stahle et al., 1992). Unfortunately, the dendrochronological record only extends to A.D. 1677 (Stahle et al., 1992). Additional dendrochronology at Reelfoot Lake may extend the tree ring record and should be undertaken to search for prehistoric earthquakes.

The Lake County uplift is a broad, low amplitude, anticline that lies within a left-stepping restraining bend in the NMSZ (Russ, 1982; Kelson et al., 1992; Schweig and Ellis, 1994). Most of the current seismicity underlies the Lake County uplift (Chiu et al., 1992). This seismicity is attributed to activity along a southwest-dipping reverse fault that underlies the Lake County uplift (Chiu et al., 1992) and reaches the surface at the base of the Reelfoot scarp (Kelson et al., 1996).

The Reelfoot scarp lies along the eastern margin of the Lake County uplift, is locally 8 m high, and extends from the southwestern margin of Reelfoot Lake to New Madrid (Van Arsdale et al., 1995a; Kelson et al., 1996) (Fig. 7). The scarp is an east-facing monocline that has recently been interpreted as the eastern limb of a fault propagation fold (Van Arsdale et al., 1995a; Kelson et al., 1996). Trenches excavated across the Reelfoot scarp have revealed normal and reverse faults; the only surface faulting within the NMSZ. Detailed trench logs have provided information on the geometry of scarp deformation and the chronology of paleoseismic events (Russ, 1982; Kelson et al., 1992,1996). Russ (1982) recognized three faulting events on the Reelfoot scarp within the last 2000 years but was unable to date the two prehistoric events. Subsequently, Kelson et al. (1992) and (1996) presented trench evidence for an 1812 and two prehistoric faulting events within the last 2400 years. Kelson et al. (1996) date the two prehistoric earthquakes as occurring between A.D. 780 and 1000, and between A.D. 1390 and 1650. Their preferred dates are approximately A.D. 900 and A.D. 1400. Based on these three faulting events within the last 2400 years, the Reelfoot fault is estimated to have an earthquake recurrence interval of 450 ± 50 years (Kelson et al., 1996).

2.2.5.3. Northern New Madrid seismic zone. Thus far, the northern NMSZ has been the least thoroughly investigated part of the seismic zone in terms of paleoseismology. Three sites have provided evidence of pre-1811 liquefaction. Evidence from this area was first described by Saucier (1991), who successfully used archaeology to identify and date prehistoric liquefaction features at the Towosahgy archaeological site in Missouri, located about 30 km northeast of Reelfoot Lake (S in Fig. 4). He attributed these features to two pre-1811 events, one between A.D. 539 and A.D. 911, and the other about 100 years before A.D. 539. Another site (W on Fig. 4) to the southwest of the Towosahgy site contains two sand blows, one likely from the 1811–1812 earthquakes, the other has a soil developed on it that dates from A.D. 770 to A.D. 1020 (Li et al., 1994). The northernmost site in the region is marked N on Fig. 4. It predates A.D. 1620 (Li et al., 1994).

2.2.5.4. Summary of paleoseismological studies in the New Madrid seismic zone. The studies discussed above all suggest that 1811 was not the first time in the Holocene that the NMSZ experienced strong ground shaking. In fact, the data are all consistent with at least three earthquakes in the 2000 years prior to 1811. However, the magnitudes of the causative earthquakes are difficult to determine. The only known post-1812 earthquake in the New Madrid region large enough to have caused liquefaction is the 1895 Charleston, Missouri, earthquake with a body-wave magnitude (m_b) of 6.2 or moment magnitude (**M**) of 6.8 (Hamilton and Johnston, 1990), which caused liquefaction over an area 16 km across (Obermeier, 1988). Saucier used empirical relationships between earthquake magnitude and maximum distance from the source of significant liquefaction (Youd and Perkins, 1978) to conservatively estimate $m_b = 6.2$ as the minimum-magnitude earthquake that could be responsible for liquefaction at both the Towosahgy and Reelfoot scarp sites, separated by a distance of about 35 km. Schweig and Ellis (1994) estimated that, when the severity

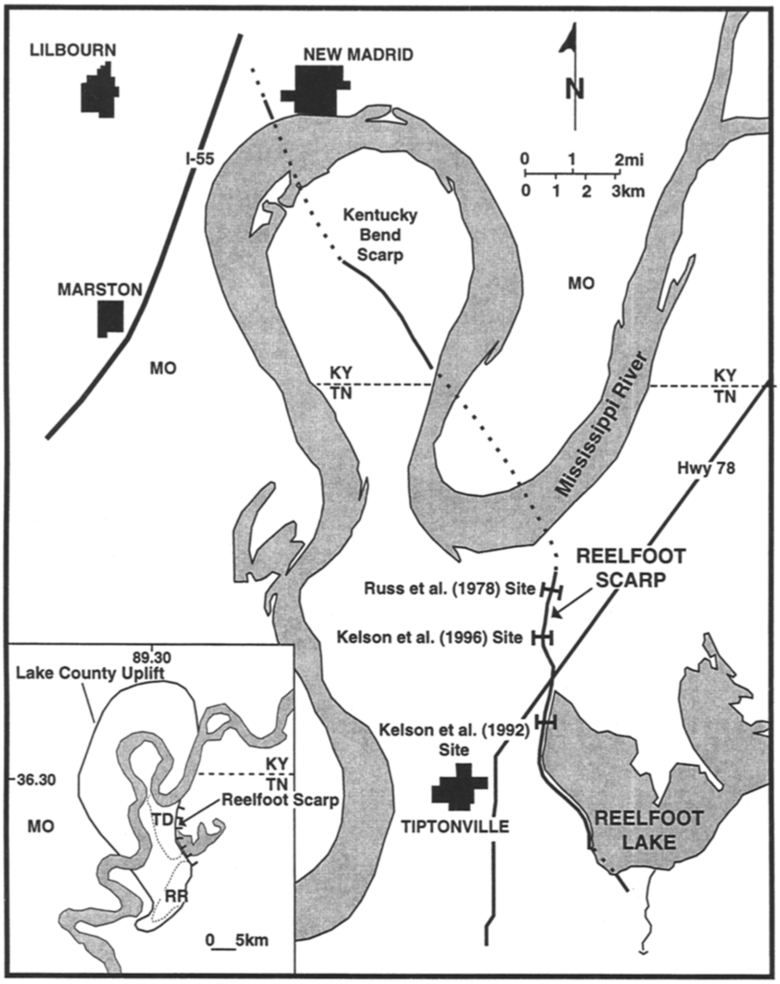

Fig. 7. Area of Reelfoot scarp and its northern extension into Kentucky (modified from Van Arsdale et al., 1995a). Trench sites also shown. Abbreviations in inset: TD, Tiptonville dome; RR, Ridgely ridge.

of liquefaction is taken into account (Youd et al., 1989), a moment magnitude 8 earthquake would be required for an event to have caused the liquefaction at sites S and E, separated by about 100 km. Much work remains to be done to complete a Holocene earthquake chronology for the region, but the results thus far seem to be in agreement with the high strain rates suggested by seismology and geodetics, as discussed previously.

3. Discussion and conclusions

The relatively high rate of neotectonic activity in the Mississippi embayment would seem to be unique in the North American stable continental interior or in stable plate interiors in general where large earthquakes are rare and surface manifestations of these earthquakes are even rarer. It is important to realize, however, that we only have access to a snapshot in time. The historical record only stretches back a couple of centuries and the instrumental record a couple of decades. The importance of this is underscored by the following two examples:

First, the Meers fault of Oklahoma is in the stable continental interior and also associated with Late Precambrian–Early Cambrian rifting. Unlike the New Madrid region, surface faulting is well expressed in the topography of the Meers fault region. It has been demonstrated that there have been at least two Holocene surface faulting events with the most recent being about 1200–1300 years ago (Crone and Luza, 1990; Kelson and Swan, 1990). Yet, the fault is aseismic today and no significant historical earthquakes have been associated with it. In fact, geological evidence suggests that the long-term recurrence interval may be 100 000 years or more (Crone and Luza, 1990). Thus, at least in the case of the Meers fault, the Holocene record gives a different picture than the long-term record or the instrumental record.

The second example is in Australia. Thirty years ago, it could have been stated that no earthquakes large enough to cause ground rupture had occurred in the stable continental interior of Australia in historic time. Since then five large earthquakes have caused ground rupture on that same continent (e.g., Crone et al., 1992). This is particularly extraordinary when one realizes that only 11 historic earthquakes have caused ground rupture in stable continental regions worldwide (Johnston, 1994). Crone et al. (1992) and Machette et al. (1993) examined two of these ruptures and found no clear evidence for earlier activity on the causative faults since Precambrian time and concluded that the recurrence intervals on these faults must be hundreds of thousands of years. Again, the Australian example indicates that one modern time sample (say the past 30 years) can give a very different view of seismic hazards in a stable continental interior from another modern time sample (say the preceding 30 years).

These two examples point out the danger of assuming that the present rate of activity correlates with past rates. Nonetheless, clearly earthquakes in the New Madrid region occurred more than one time in the Holocene, as shown in the geological record. Yet the low topographic and structural relief in the region does suggest that tectonic activity, at least in its present incarnation, cannot have been going on very long, perhaps no more than a few tens of thousands of years (Pratt, 1994; Schweig and Ellis, 1994). The evidence for Pleistocene or younger faulting in the Benton Hills and liquefaction in the Western Lowlands suggests that neotectonic activity may jump around the Mississippi embayment with time.

No satisfactory explanation for the relatively high rates of neotectonic activity in the upper Mississippi embayment have been suggested. The one major structural feature associated with the activity is the Reelfoot rift (e.g., Braile et al., 1982,1984; Hildenbrand et al., 1982), which is favorably oriented with respect to regional stresses. Although this, in itself, is not unique in stable North America, it has been suggested that the rift in combination with other features may be the cause of the intense earthquake activity in the region. These features include: (1) igneous bodies that may cause stress concentrations due to their rheological contrast with the host rocks (e.g., Hildenbrand and Hendricks, 1995; Hildenbrand et al., 1995; McKeown, 1982; Ravat et al., 1987); the intersection of the rift with a major northwest-trending gravity anomaly, the Missouri gravity low (Hildenbrand and Hendricks, 1995) (this intersection may be a zone of crustal weakness); high pore fluid pressures (McKeown and Diehl, 1994; Zoback and Zoback, 1992), perhaps associated with the Blytheville arch, a 5–10-km wide structural high within the axis of the rift (Hamilton and McKeown, 1988).

The upper Mississippi embayment has a long and varied tectonic history. Holocene and current tectonic activity seems to be at a much higher rate than for other areas of stable continental interiors, yet the cause remains elusive. Ongoing geodetic and geological studies should provide more insight into the precise manner in which crustal strain is accumulating, and perhaps will allow improved regional neotectonic models.

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